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Surface Climate and Snow–Weather Relationships of the Kuparuk Basin on Alaska’s Arctic Slope

Peter Q. Olsson, Larry D. Hinzman, Matthew Sturm,
Glen E. Liston, and Douglas L. Kane

August 2002

Abstract: This report summarizes temperature, wind, and snow-cover data for the Kuparuk River Basin in Arctic Alaska spanning the five-year period of 1994–1998. Comparison of results from five meteorological towers is presented to illustrate both the differences and similarities of the regional climate and weather along a 200-km transect. A picture emerges of the Arctic Slope as a region dominated by subfreezing temperatures for most of the annual cycle. The five sites showed a good deal of similarity in both seasonal variation and meteorological forcing on a time scale of a few days. While the Kuparuk Basin is typified by al-

most constant moderate winds, winds greater than 10 m s^{-1} are fairly rare. The observed patterns of temperature and wind have important ramifications for the winter snow cover of the Kuparuk Basin, explaining why the snow cover forms first in the foothills and last near the coast. The surprisingly low wind speeds across the network in October and November help to explain the presence of thick but low-density layers observed in the basal snow-pack. A pronounced warming event occurring each November capped this early snow with a melt crust or wind slab or both, protecting it from subsequent wind erosion.

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PREFACE

This report was prepared by Peter Q. Olsson, Chief Scientist, Alaska Experimental Forecast Facility, University of Alaska Anchorage; Larry D. Hinzman, Professor, Water and Environmental Research Center, University of Alaska Fairbanks; Matthew Sturm, Research Physical Scientist, U.S. Army Cold Regions Research and Engineering Laboratory, Fairbanks, Alaska; Glen E. Liston, Research Scientist, Department of Atmospheric Sciences, Colorado State University; and Douglas L. Kane, Director, Water and Environmental Research Center, University of Alaska Fairbanks.

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CONTENTS

Preface	ii
Contents	iii
1 Introduction and Background	1
2 Site Descriptions	5
3 Meteorological Instrumentation	7
4 Results	9
Temperature Data Averaging	9
Temperatures	10
Winds	17
5 Snow Cover	27
6 Discussion	31
7 Summary and Conclusions	36
Literature Cited	38

LIST OF ILLUSTRATIONS

Figure 1. Locations of the five surface meteorological sites located in the Kugaruk Basin used in this study	3
Figure 2. Typical meteorological tower used in this study	7
Figure 3. Eleven-day running average of the mean daily temperature for the five sites over the annual cycle	10
Figure 4. Monthly mean temperatures for the five-year period for each of the sites	11
Figure 5. Eleven-day running averages of maximum and minimum daily temperatures, and diurnal temperature range for the four seasonal periods.	12
Figure 6. Eleven-day running averages of daily mean wind speed, average maximum wind speed, and dispersion for each day for the four seasonal periods.	21
Figure 7. Seasonal wind rose diagrams for the five sites and for Prudhoe Bay using three threshold minimum velocities	22
Figure 8. Line of the winter traverses in the Kugaruk Basin and the locations of the seven snow pits shown in Figure 9.	27
Figure 9. Stratigraphy for the Kugaruk Basin, winter of 1996-97	28
Figure 10. Snow stratigraphy from the Kugaruk Basin for the winters of 1994-95 through 1997-98	30
Figure 11. Daily mean temperatures at the five sites for Oct. 1997 through Jan. 1998	31
Figure 12. Daily mean temperatures at SAG for all five years of record during the latter half of <i>EC</i>	33
Figure 13. Synoptic conditions for 12 UTC, 8 Nov. 1997	34

LIST OF TABLES

Table 1. Acronyms used in this report.	9
Table 2. Record maximum and minimum temperatures and mean temperature for each month and site for the period of record	13
Table 3. Mean, earliest, and latest dates at each site for the occurrence of temperatures less than 0°, -3.0°, and -6.0°C lasting for at least six consecutive hours.	15
Table 4. Monthly mean and record maximum wind speeds for each month.	19

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1 INTRODUCTION AND BACKGROUND

The Arctic Slope of Alaska is generally defined as that portion of Alaska north of the crest of the Brooks Range. This region, with an area over 230,000 km², is one of the least-understood climatic regions of the United States and is certainly the most sparsely populated. At the same time, it is estimated that more than 30% of the nation’s domestic petroleum is produced there. Economic development and the concomitant concern over its potential environmental impact have been the impetus for much of the geophysical research that has been conducted on the Arctic Slope to date. Along with renewed political and economic interest in petroleum exploration of the Arctic National Wildlife Refuge (ANWR) in the eastern part of the Arctic Slope comes a heightened concern for the integrity of the fragile ecosystem of this region in the face of a significant anthropogenic disturbance.

High-latitude climates are thought to be particularly sensitive and variable. Recent studies in this area (Kattenburg et al. 1996, Oechel et al. 2000, Sturm et al. 2001) show a tight coupling between the biosphere and the physical environment, suggesting that both natural and anthropogenic changes in climate and the planetary surface would be expected to induce responses in the biota inhabiting the region. A more complete understanding of the current Arctic Slope climate is necessary to quantify change as it occurs and to anticipate and potentially mitigate the effects induced by alterations of the geophysical environment. This report attempts to quantify a critical subset of the full Arctic Slope climate system as it is today—namely the climate of the cold season, which we define here to be that period dominated by subfreezing temperatures and snow cover.

Our focus on the cold season is motivated by the following points:

- The cold season lasts from September to June and is the norm for the Arctic, occupying nearly $\frac{3}{4}$ of the annual cycle.

- One of the most important outcomes of the winter climate of the Arctic Slope is the creation of a snow cover. This snow cover protects and insulates the ground and low-lying plants, reduces desiccation, and maintains ground temperatures that range from 5° to 30°C higher than air temperatures. Active-layer thickness and spring snowmelt run-off, often the peak discharge of the year, all depend on the amount and distribution of the snow. In addition, the formation of a snow cover protects the fragile tundra from disturbance by human activity such as seismic exploration.
- It has become increasingly apparent that biological processes with climatic significance do not completely cease during the cold season. For example, microbial decomposition and CO₂ production continue well after the cold season has started (Clein and Schimel 1995, Fahnestock et al. 1998, Zimov et al. 1993).
- One of the most obvious Arctic climate feedbacks results from the differing albedoes of snow-covered vs. bare tundra. The timing of snowfall events and the subsequent redistribution and weathering of snow by the wind determines the small-scale “patchiness” observed during the thaw. This mosaic of bare and snow-covered tundra has a significant effect on local radiation budgets (Dingman et al. 1980, Kane et al. 1991, Liston 1995).
- The cold season has a large impact on humans living and working on the Arctic Slope. As the surface freezes and the snow cover develops, overland transportation becomes easier and less destructive because the tundra is covered.

As is the case with much of Alaska, Arctic Slope communities and facilities with a significant period of climate record are located either on the coast (e.g. Barrow, Prudhoe Bay, and Wainwright), on islands (e.g. Barter Island), or in immediate proximity to rivers (e.g., Atkasak, Umiat, and Nuiqsit). The National Weather Service operates only five permanent weather stations on the Arctic Slope, four of which are located within a few kilometers of the coast. Analysis of these records tends to bias the result, rather than giving a picture of Arctic Slope climate as a whole. This bias is compounded by the ubiquitous near-surface temperature inversions commonly found at high latitudes during the Arctic winter. The accompanying thermal stability strongly inhibits vertical mixing of momentum and makes possible an extremely variable local microclimate that can bias the climate record. This, in combination with terrain influences, can produce local conditions, particularly wind direction, that are in direct opposition to large-scale patterns (e.g., Haugen et al. 1976, Schwerdfeger 1973).

Most studies of Arctic Slope climate have relied almost exclusively on the coastal and coastal plain records (e.g., Searby and Hunter 1971, Wiseman and Short 1976) or have concentrated on specific locations (Clebsch and Shanks 1968, Brown et al. 1975, Haugen et al. 1976, Thoman 1995, Wendler 1978). Researchers have also used analyses of synoptic upper air data to derive synoptic climatologies of the Arctic Slope (e.g., Moritz 1979, Milkovich 1991) and mean cyclone tracks in the Alaskan Arctic (Serreze et al. 1993).

Some studies have attempted to gain a better understanding of the more inland reaches of the coastal plain and adjacent uplands, but this work has been limited to the warm season. Walsh (1977) used aircraft data to gain a more integrated view of the lower summer troposphere along the northern coast of Alaska. Haugen and Brown (1980) used a regression technique on data from stations on the inland tundra as well as the coastal locations to estimate mean temperatures across the coastal plain as a function of longitude as well as latitude.

In this report we summarize five years of continuous surface meteorological data taken along a transect from near Prudhoe Bay on the Arctic Coast to the foothills of the Brooks Range (Kane and Hinzman 2000) that includes stations in the inland reaches of the Arctic Slope (Fig. 1). The climate data from the meteorological sites are augmented by snowpack surveys conducted several times each winter during the study period. To our knowledge this study represents the only surface climatic analysis that has been conducted of this region since

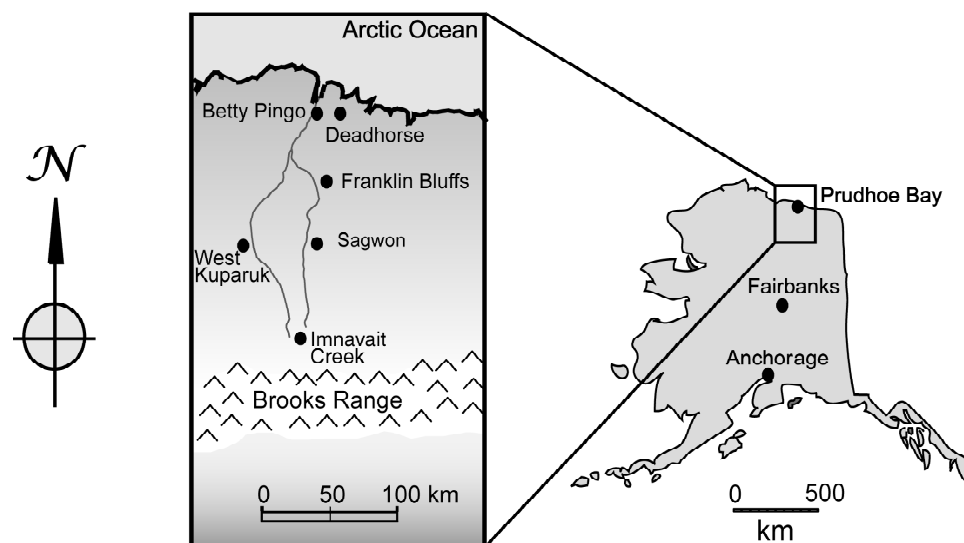


Figure 1. Locations of the five surface meteorological sites located in the Kuparuk Basin used in this study: Betty Pingo (BET), Franklin Bluffs (FRA), Sagwon (SAG), West Kuparuk (WKU), and Imnavait Creek (IMN).

Haugen (1980, 1982). Through our emphasis on the cold season, we also quantify several points that are considered common knowledge among Arctic inhabitants yet may seem counterintuitive to persons inhabiting lower latitudes. Additionally this work can serve as a baseline against which future climatic trends may be detected.

This is also the first study to emphasize contemporary multi-year cold season measurements in upland as well as coastal plain locations and to include both wind speed and direction measurements throughout the period of record. Although not discussed here, we recognize that precipitation is also an important aspect of the winter climate. While some precipitation measurements are available for this region, the automated devices used were not designed to accurately measure the solid-phase precipitation occurring throughout the cold season. Consistent measurements of precipitation throughout the annual cycle are especially difficult to achieve in the Arctic environment (e.g., Black 1954, Benson 1982, Goodison et al. 1985, Yang et al. 1999).

In the following section we describe the siting of the various stations comprising the tower network and the meteorological instrumentation deployed at these sites, and we discuss issues associated with the data derived from the network. Next, we present the results of the data analysis and ancillary snow cover measurements, followed by a discussion of the results and some specific weather issues illuminated by the climate analysis. Finally we summarize the material presented in the context of weather/snow cover interactions and climate change research.

2 SITE DESCRIPTIONS

Starting in 1985 on Alaska's Arctic Slope, meteorological stations at various research sites were located in a roughly south–north transect along the Dalton Highway (Fig. 1). All of the sites are within or near the Kuparuk River basin.

The southernmost site (IMN) was established in the foothills of the Philip Smith Mountains near the headwaters of Imnavait Creek (latitude 68°37'N, longitude 149°18'W, 940 m elevation). Upstream of the gaging site, this watershed (drainage area 2.2 km²) has an average elevation of 910 m and is located about 165 km south of Prudhoe Bay and 15 km north of the Brooks Range. Instrumentation for collecting meteorologic, hydrologic, and soil data was installed in 1984, and data collection began in the spring of 1985. The mineral soils in this area are cold, wet, poorly drained silt loams with a high organic content and include many glacial erratics of various sizes. The mineral soils are covered by a peaty layer and are classified as Histic Pergelic Cryaquepts (Rieger et al. 1979). The vegetation is mostly water-tolerant plants such as tussock sedges and mosses, but there are also lichens and shrubs such as willows, alders, and dwarf birches. More complete descriptions of tundra vegetation have been published (Brown and Berg 1980, Walker et al. 1989). The area was glaciated during the Pleistocene and is underlain by continuous permafrost. The maximum depth of the active layer averaged 53 cm between 1992 and 1999 (Brown and Hinkel 2000).

In 1986, two additional meteorological sites were established north of the Imnavait Creek site. For access reasons, both sites were located along the Dalton Highway, just outside the Kuparuk River watershed boundary. One site (SAG) was established on a hilltop near the Sagwon Bluffs (latitude 69°25'N, longitude 148°42'W) approximately 100 km south of Prudhoe Bay. This site is located in a transitional zone between the coastal plain and the foothills at an elevation of 300 m. The vegetation is also characteristic of tussock tundra, and the soils are loamy with a peaty surface layer and are poorly drained (Everett 1980). Instrumentation for measuring soil temperatures and meteorologic conditions was installed near the top of a 10% north-facing slope. The depth of thaw above permafrost is typically about 50 cm.

Also in 1986 a meteorological research area (FRA) near Franklin Bluffs (latitude 69°54'N, longitude 148°46'W) was established on the coastal plain 50 km south of Prudhoe Bay. This site is located in the relatively flat area of the Sagavanirktok River floodplain at an elevation of 80 m. The vegetation consists of a continuous cover of grasses and sedges rooted in mosses and lichens (Komarkova and Webber 1980). The soils are poorly drained and generally do

not thaw to depths of more than 50 cm. Organic materials of variable thickness overlie silt-loam-textured mineral soils (Everett 1980). The maximum depth of soil thawed above the permafrost averaged 66 cm between 1996 and 1999 (Brown and Hinkel 2000).

A wetland study site (BET) (latitude 70°17'N, longitude 148°54'W) was established near Betty Pingo within the Prudhoe Bay oilfield in May 1994. This site was located on the Arctic Coastal Plain in an area of little topographic relief at an elevation of about 12 m and about 12 km from the coast. The vegetation consists of wet sedge tundra and forb tundra. The soils are organic overlying layers of fine sand and silts. The 10-m tower was established in a wetland area; however, nearby dryer areas were also equipped with soil thermistors and net radiation instrumentation. The maximum depth of soil thawed above the permafrost averaged 55 cm between 1993 and 1999 (Brown and Hinkel 2000).

In July 1994 the West Kuparuk site (WKU) (latitude 69°26'N, longitude 150°20'W) was established near the western margin of the Kuparuk Basin approximately 63 km west of the Sagwon site. The site is located on a hill overlooking the main branch of the Kuparuk River at an elevation of about 160 m. The vegetation is characteristic of tussock tundra. The soils, which are loamy with a peaty surface layer and are poorly drained, are very similar to the Sagwon site. The depth of thaw above permafrost is typically about 50 cm.

Two other major meteorological sites exist in the Kuparuk River basin: West Dock and Upper Kuparuk River. The West Dock site was established along the Arctic coastline near the Betty Pingo site where strong horizontal gradients of temperature and wind speed exist. The Upper Kuparuk River site is about 5 km from the Imnavait Creek site, where meteorological variation exists over vertical gradients of 170 m. Five micrometeorological sites exist in the headwaters where summer precipitation and continuous air temperature, relative humidity, and wind speed are measured.

3 METEOROLOGICAL INSTRUMENTATION

Campbell Scientific CR10X dataloggers were used to record and process data at all sites. Sensors were sampled once per minute and averaged hourly. Data recorded on the dataloggers were compared to measured conditions to check the sensor calibrations and the datalogger during site visits. A 10-m meteorological tower was used to mount the air temperature, relative humidity, wind speed, and wind direction sensors (Fig. 2). Radiation sensors were mounted on a separate tower designed to suspend these sensors over the tundra and minimize shadows.

Wind speed was measured using Met-One model 014A anemometers at 1-, 3-, and 10-m heights. The threshold velocity of this instrument is rated at 0.45 m s^{-1} ,

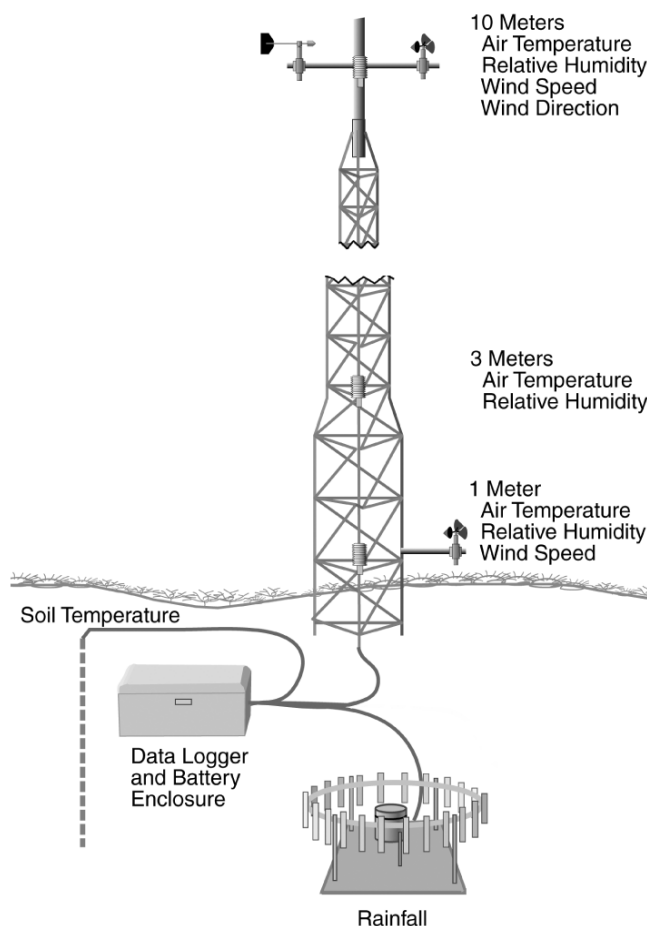


Figure 2. Typical meteorological tower used in this study.

and the reported accuracy is approximately 0.11 m s^{-1} . Wind direction was measured with a model 024a Met-One wind direction sensor. The accuracy, as reported by the manufacturer, is within $\pm 5^\circ$. The wind direction sensors are calibrated annually by obtaining the full-scale return of the resistor and scaling this to the desired 360° output. Additionally, the heading of the wind direction sensors are checked periodically each year by pointing the vane at four compass points.

Radiation instruments are annually installed in the spring, usually during April, and are taken down in the fall, typically early September. Since rime ice, nocturnal frost, snowfall, and freezing precipitation can obscure the sensors in these instruments, values reported during periods of below-freezing air temperature must be used cautiously. The following radiation components were measured: incoming and reflected short-wave radiation, atmospheric and terrestrial long-wave radiation, and net radiation. Radiometer calibrations were checked locally each year by comparison to the output of an instrument of known precision. All radiometers are frequently sent to independent laboratories for reconditioning and recalibration.

Incident short-wave radiation was measured using an Eppley model PSP precision spectral pyranometer. This instrument has a reported spectral range of $0.285\text{--}2.800 \text{ }\mu\text{m}$ and a reported accuracy of $\pm 1\%$ in the range of values encountered. The cosine response of this instrument is $\pm 1\%$ between 0° and 70° and $\pm 3\%$ between 70° and 80° zenith angle. Eppley model PIR precision infrared pyrgeometers were used to measure long-wave radiation, both terrestrial and atmospheric. The spectral range of this type of instrument is $4\text{--}50 \text{ }\mu\text{m}$, and the accuracy reported as $\pm 1\%$ between 0 and 700 W m^{-2} . Net radiation was measured with Q-7.1 sensors manufactured by Radiation and Energy Systems. The spectral range of this type of instrument is $0.25\text{--}60 \text{ }\mu\text{m}$, and the accuracy is not reported.

Air temperature and relative humidity were measured throughout the year with a Campbell Scientific HMP35C or HMP45C. The accuracy of the air temperature is within 0.45° across the measurement range. The accuracy of the relative humidity sensor is reported to be within 3% at worst case. Warm-season rainfall precipitation is measured with a TE525 tipping bucket rain gage manufactured by Texas Electronics. The rain gage is calibrated to record 0.01 in. of rain per tip. In the low-intensity rain events typically observed (less than 1 in./hr) the accuracy is reported at $\pm 1\%$. No precipitation measurements were taken during the cold season.

This report only considers a small subset of the data that were collected as part of this ongoing effort (Kane et al. 2000). The complete data archive can be found at the National Snow and Ice Data Center in Boulder, Colorado (<http://nsidc.org/data/arcss015.html>).

4 RESULTS

In discussions of the annual climate cycle, it is common to decompose the yearly cycle into three-month seasons defined on monthly boundaries rather than the traditional seasonal definitions based on the solar cycle. We follow the former course here, using the monthly groupings of Sep.–Oct.–Nov., Dec.–Jan.–Feb, Mar.–Apr.–May, and Jun.–Jul.–Aug and further naming these seasons Early Cold (*EC*), Deep Cold (*DC*), Late Cold (*LC*), and Warm Season (*WS*), respectively. (Refer to Table 1 for a summary of acronyms.) Our rationale for this nomenclature is based on the annual temperature cycle and will be discussed further as that cycle is considered.

Table 1. Acronyms used in this report.

BET	Betty Pingo	EC	Early Cold
FRA	Franklin Bluffs	DC	Deep Cold
SAG	Sagwon	LC	Late Cold
WKU	West Kuparuk	WS	Warm Season
IMN	Imnavait Creek		

Temperature Data Averaging

Climate data typically show a great deal of temporal variability on scales ranging from diurnal to interannual to millennial. The five-year period of record (1994–1998) discussed here is clearly insufficient to address those time scales (decadal and upward) relevant to what is heuristically referred to as climate change. Rather, our intent here is to quantify the current climate along the Arctic Slope as well as possible with the limits of the data record available to us.

The five-year period of record used here shows a considerable degree of interannual variability for any given day of year, and the limited number of samples (at best five, assuming no missing data points) permit extreme short-term excursions from a mean state to produce highly variable (i.e., noisy) time series of such things as mean daily temperature when averaged over the period of record. Seasonal trends, a central focus of this work, are more apparent after some form of smoothing is applied to the data. Much of the data presented are in the form of running averages. After experimenting with different averaging periods, we found that an averaging period of 11 days was sufficient to inhibit the short-term variability associated with propagating weather disturbances while permitting the best possible resolution of evolving seasonal trends. Clearly, if the period of record were several times longer, such filtering would become less

important. With only five years of data, however, individual synoptic events with a time scale of a few days sometimes dominate the unfiltered record.

Temperatures

A running average of the mean daily temperature for all five sites for the annual cycle is shown in Figure 3, along with a graphical depiction of our seasonal nomenclature. Figure 4 shows monthly mean temperatures for the sites. Figure 5 shows running averages of daily maximum and minimum temperatures and diurnal temperature ranges by season. The rationale for the season names becomes apparent when considering the location of the freezing isotherm in the daily minimum temperature plot.

Warm Season

The Warm Season (*WS*) corresponds closely with that period of the year when diurnal mean temperatures are consistently above freezing (Fig. 5) and active plant growth occurs. At most sites the warmest temperatures of the year occur in early to mid-July, just a few weeks after the summer solstice, when daily

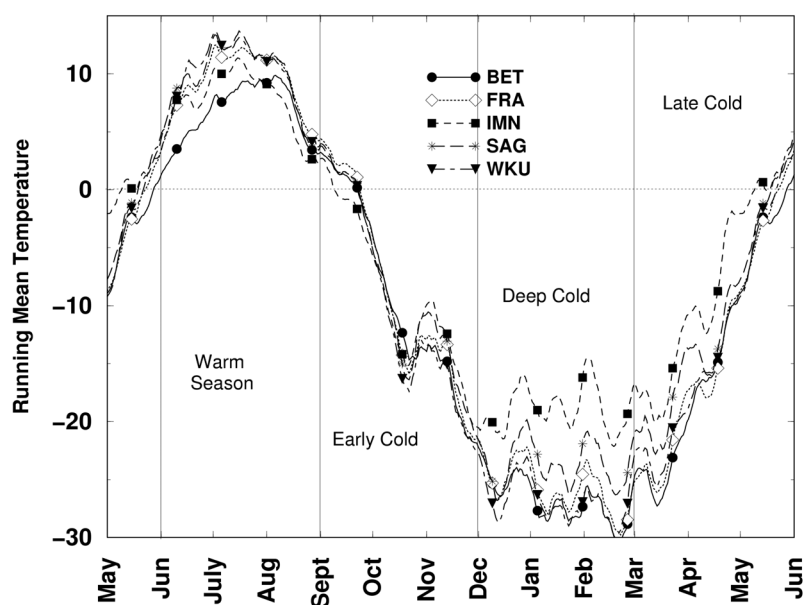


Figure 3. Eleven-day running average of the mean daily temperature for the five sites over the annual cycle. The axis label corresponds to the first day of that month. Also graphically depicted are the seasonal definitions for Warm Season (*WS*), Early Cold (*EC*), Deep Cold (*DC*), and Late Cold (*LC*).

solar forcing is the largest. The near-coastal site BET is an exception, with both maximum and minimum daily temperatures lagging several degrees behind the inland sites. The warmest temperatures at BET are found in late July into early August, when the sea ice has typically retreated from the Beaufort Coast and solar heating is still a significant contributor to the radiation budget. The low maximum temperatures and small diurnal temperature variations at IMN in July and August are a consequence of its location in the foothills of the Brooks Range, where, in addition to increased elevation, almost daily convective activity results in cloud cover during much of the late-summer heating cycle.

Table 2 contains monthly mean, record maximum, and record minimum temperatures for each month. (The record temperatures are, of course, the extrema for only the five-year period considered here.) While not the warmest location in terms of monthly means, FRA, located on the inland side of the coastal plain, had the warmest July and August record temperatures by far. The record minima also reveal that all locations also recorded sub-freezing temperatures at least once during each month of *WS*.

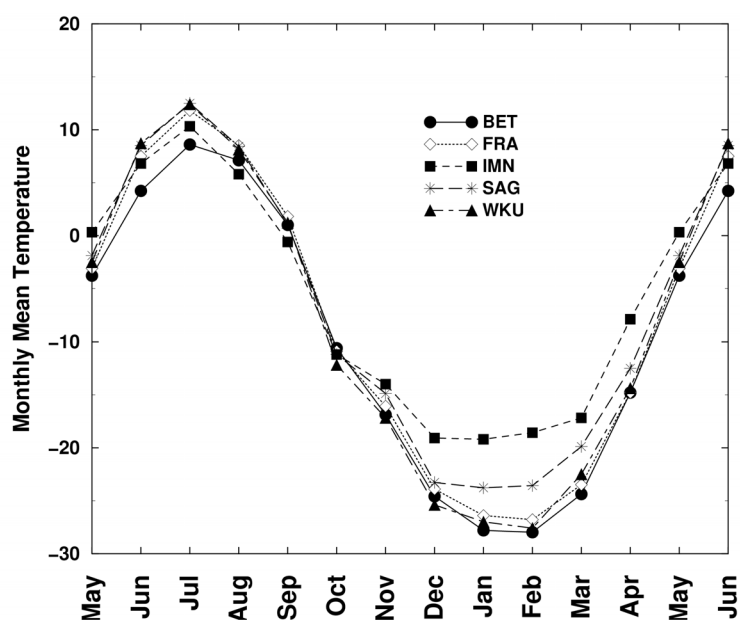


Figure 4. Monthly mean temperatures for the five-year period for each of the sites.

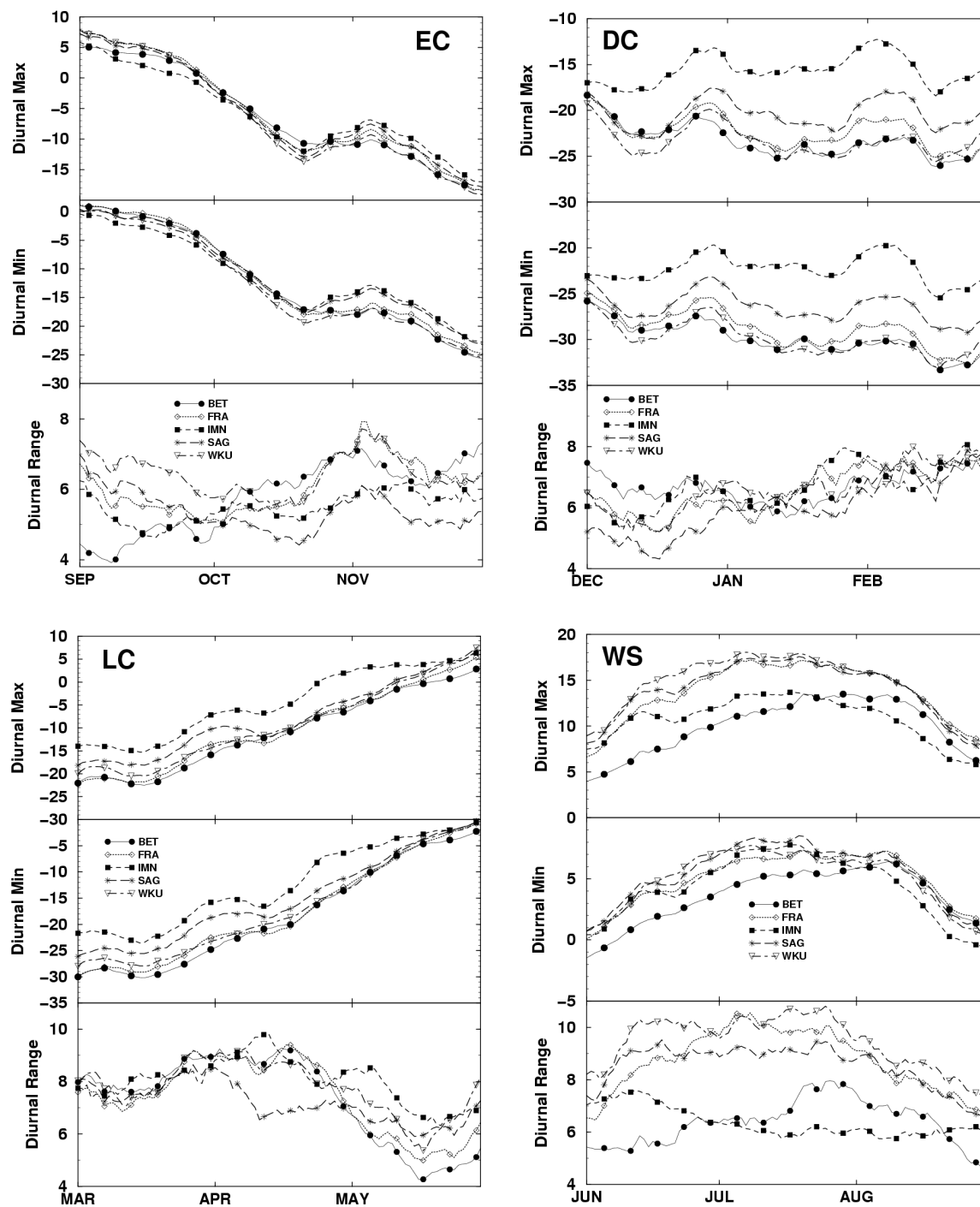


Figure 5. Eleven-day running averages of maximum and minimum daily temperatures, and diurnal temperature range for the four seasonal periods.

Table 2. Record maximum and minimum temperatures and mean temperature for each month and site for the period of record. The day of month and the year when the temperature occurred are also given.

		BET		FRA		IMN		SAG		WKU	
		Temp. (°C)	Day- year	Temp. (°C)	Day- year	Temp. (°C)	Day- year	Temp. (°C)	Day- year	Temp. (°C)	Day- year
EC											
Sept.	Record max.	18.3	21-95	22.4	22-95	17.6	18-95	22.6	21-95	21.2	18-95
	Record min.	−17.6	28-96	−18.6	28-96	−18.5	27-96	−15.0	27-96	−17.8	28-96
	Monthly mean	1.0		1.8		−0.6		1.1		1.2	
Oct.	Record max.	1.7	12-98	3.8	30-95	5.8	09-98	3.7	31-95	3.4	30-95
	Record min.	−34.0	25-96	−31.2	24-96	−35.2	24-96	−31.0	24-96	−31.1	24-96
	Monthly mean	−10.6		−10.9		−11.2		−11.2		−12.2	
Nov.	Record max.	3.2	11-97	5.7	11-97	4.2	01-95	8.0	10-97	4.6	10-97
	Record min.	−35.2	30-94	−34.6	22-95	−33.2	21-94	−33.1	21-94	−35.4	07-94
	Monthly mean	−16.9		−16.1		−14.0		−14.9		−17.2	
DC											
Dec.	Record max.	0.5	19-98	2.1	19-98	0.8	19-98	1.4	19-98	1.1	19-98
	Record min.	−42.0	22-97	−42.5	21-97	−39.7	19-97	−40.1	21-97	−41.9	18-97
	Monthly mean	−24.6		−23.9		−19.1		−23.3		−25.4	
Jan.	Record max.	−3.2	29-96	−0.1	29-96	2.0	06-95	1.2	02-95	0.6	29-96
	Record min.	−42.8	30-95	−42.8	21-97	−39.6	17-98	−40.2	16-98	−42.8	29-96
	Monthly mean	−27.8		−26.4		−19.1		−24.1		−27.3	
Feb.	Record max.	−5.4	12-95	−2.5	02-94	3.9	01-94	3.4	02-94	−3.6	25-96
	Record min.	−42.8	23-95	−46.1	27-97	−40.8	05-96	−41.7	18-98	−44.0	18-98
	Monthly mean	−28.0		−26.8		−18.6		−23.6		−27.6	

Table 2 (cont.). Record maximum and minimum temperatures and mean temperature for each month and site for the period of record.											
		BET		FRA		IMN		SAG		WKU	
		Temp. (°C)	Day- year	Temp. (°C)	Day- year	Temp. (°C)	Day- year	Temp. (°C)	Day- year	Temp. (°C)	Day- year
LC											
Mar.	Record max.	-1.7	26-96	3.4	21-98	3.9	20-98	5.2	20-98	4.0	21-98
	Record min.	-44.6	12-95	-45.3	14-95	-42.8	01-94	-42.0	15-95	-43.7	15-95
	Monthly mean	-24.4		-23.5		-17.2		-19.9		-22.5	
Apr.	Record max.	4.5	15-98	8.2	25-95	10.5	30-95	10.8	30-95	6.5	25-95
	Record min.	-38.4	02-97	-38.7	02-97	-33.2	12-94	-35.0	02-97	-38.4	02-97
	Monthly mean	-14.8		-14.8		-7.9		-12.5		-14.4	
May	Record max.	11.6	27-98	15.9	24-96	16.9	12-95	17.7	28-98	17.8	29-96
	Record min.	-26.4	03-96	-24.9	02-96	-18.6	02-96	-22.8	05-98	-26.5	02-96
	Monthly mean	-3.8		-3.1		0.3		-1.9		-2.5	
WS											
Jun.	Record max.	24.8	29-98	25.8	29-98	19.4	16-96	24.0	15-96	24.4	29-98
	Record min.	-9.9	02-97	-9.2	02-97	-9.2	02-97	-10.5	02-97	-8.1	03-97
	Monthly mean	4.2		7.5		6.8		8.5		8.7	
Jul.	Record max.	25.8	02-94	32.6	29-94	21.5	07-95	28.7	29-94	26.7	01-98
	Record min.	-1.3	09-96	-2.3	10-96	-2.5	25-95	-3.4	09-96	-2.1	09-96
	Monthly mean	8.6		11.8		10.3		12.5		12.4	
Aug.	Record max.	27.5	01-94	32.3	01-94	21.9	06-94	27.7	01-94	27.3	06-94
	Record min.	-3.5	28-96	-4.5	23-96	-9.9	26-96	-6.8	22-96	-6.0	28-96
	Monthly mean	7.1		8.5		5.8		8.4		8.1	

In contrast to the uplands, late *WS* into early *EC* is typically the cloudiest time of year on the coast, and it is at this time that BET sees its smallest diurnal variation in temperature. It is curious to note that by the *end* of *EC*, BET has the *highest* diurnal range of all sites. The brief warming trend seen in almost all locations starting in late October is another interesting feature in an otherwise monotonically decreasing temperature trend at all sites. (This will be considered further in the discussion section.) It is also noteworthy that the period from the last week of September through about the third week of October is when temperatures are most homogeneous over the network.

IMN is the first of the sites to reaching sub-freezing minimum temperatures, this occurring in the last third of August. By the first week in September all sites are experiencing below-freezing temperatures, and by the end of September temperatures at all sites typically remain below 0°C throughout the day. Table 3 contains date statistics for freezes (≤ 0 , -3 , -6°C for at least six consecutive hours) for each of the sites, based on a frost-free period starting on 15 July. (The notion of a truly “frost-free season” in the traditional midlatitude sense is somewhat nebulous here, given that temperatures on the Arctic Slope occasionally dip below freezing for short periods throughout *WS*. Hence, the six-hour time criterion and negative threshold temperatures give a more realistic measure of “significant frosts.”) Higher-elevation IMN is the outlying site in this measure, with mean dates typically preceding the other sites by about two weeks. The two sites closest to marine modifying influences, BET and FRA, are the latest in each category. These date trends have a significant impact on the date that snow cover begins to form.

Table 3. Mean, earliest, and latest dates at each site for the occurrence of temperatures less than 0°, –3.0°, and –6.0°C lasting for at least six consecutive hours.

	$\leq 0^\circ\text{C}$ for 6 hours			-3°C for 6 hours			-6°C for 6 hours		
	Mean	Early	Late	Mean	Early	Late	Mean	Early	Late
BET	08/25	07/28	09/19	09/18	09/04	10/03	10/02	09/16	10/13
FRA	08/28	08/10	09/12	09/12	08/21	09/30	09/29	09/13	10/10
IMN	08/09	07/25	08/21	08/24	08/10	09/11	09/05	08/22	09/22
SAG	08/21	07/29	09/27	09/08	08/11	09/29	09/20	08/23	10/05
WKU	08/22	08/10	09/06	09/05	08/21	09/22	09/22	09/15	10/04

Deep Cold

Deep Cold (*DC*) differs from *EC* in that the long-term temperature trend is more or less flat but punctuated with strong temperature fluctuations of shorter duration. Even after the averaging and filtering operations, a fair degree of short-term variability is apparent in Figure 5, attesting to the degree of variability from a mean state at this time. During this period of almost no insolation, the Arctic Slope experiences several major polar cold outbreaks, with very few contrasting warm spells. Though several sites had record monthly maximum temperatures above freezing, this season typically sees few anomalous warm spells. Essentially, it is either cold or *very* cold. Intersite variability in temperature is largest during *DC*, with coastal BET on average consistently the coldest site and IMN the warmest. There is a fair degree of coherence in the temperature oscillations among the sites, suggesting that the cold polar air mass outbreaks responsible for the low temperatures extend across most of the network of sites. With the exception of WKU, the trend in maximum, minimum, and mean temperatures also shows warmer conditions to the south, which also corresponds to warmer temperatures with increasing elevation.

The persistent low temperatures at BET, FRA, and WKU during *DC* attest to one of the signal features of cold-season climate at inland high-latitude regions—extreme variability in microclimate dominated by infrared cooling during the long and often clear winter night. Over a period of several days the resident air mass comes into thermal equilibrium with the radiatively cooled snow surface (Curry 1983). This pooling of very cold air produces a very strong temperature inversion, similar to that seen at many locations in interior Alaska (Bowling et al. 1968). Once in place, this stable thermal environment resists vertical mixing and typically persists until replaced with a more temperate air mass.

The last half of February is the coldest time of year at all sites, even though the diurnal cycle is again becoming better established as solar radiation returns to the Arctic Slope. This fact is apparent both in mean temperature and in the record low temperatures, with the latter occurring in February at three of the five sites. The record minimum temperature for the network for the period of record considered here, -46.1°C (-51°F), occurred at FRA on 27 February 1997. Also, only two sites, IMN and SAG, had February record maximum temperatures above freezing.

Late Cold

With increasing insolation the Late Cold season (*LC*) is a period of more-or-less steady warming, punctuated by strong polar cold outbreaks that often affect the entire network, especially during March. While mean temperatures are

increasing at all locations, March record low temperatures do not differ much from those in February, the result of late cold outbreaks as the global atmosphere mixes its large reservoir of “aged” polar air equatorward. Weather events at this time are dominated by the equatorward plunges of these cold and dry polar air masses, so the months of March and April are typically relatively cloud free across much of northern Alaska, with precipitating storms at a minimum.

Following the *DC* trend, higher-elevation IMN is consistently the warm spot and also tends to have the largest diurnal variation. The northernmost sites, BET and FRA, remain the coolest sites, as they continue to be influenced by the shallow but frigid polar air masses that persistently dominate the coastal region. As *LC* progresses and the length of day increases, the diurnal range in temperatures decreases across the network. IMN consistently achieves maximum temperatures above freezing by late April, almost a month ahead of the other sites, and not surprisingly the initiation of snowmelt generally occurs there first as well. Well before breakup the increasing insolation creates small regions of bare ground in wind-exposed areas, such as ridge crests, that had thin seasonal snow cover. Generally, however, it is not until *minimum* temperatures consistently remain above freezing that the breakup of the snowpack begins in earnest (Kane et al. 1997), usually occurring at any given location in about a two-week period. The running average procedure used in Figure 5 masks this rather episodic behavior that occurs first at the most inland sites during mid-May to early June and progressively later as one moves toward the coast. After breakup has occurred, inter-site temperature differences along the network are reduced.

Winds

In general, winds are a much more difficult variable than temperature to measure and quantify. This is especially true in the Arctic cold season, where physical conditions (e.g. riming and icing) adversely impact the mechanical wind instrumentation. Hence, wind data from these sites tend to be less robust than the temperature record. Additional factors, such as curious bears and other wildlife, create problems in the warm season. Since the sites used in this study are unmanned for all but a few brief visits per year, instrument performance can be degraded for long periods of time. Fortunately winds were measured at several heights on each tower, providing some redundancy in measurement. Whenever possible, wind measurements at 3 m were used, with a correction factor based on a logarithmic profile assumption for each site being applied to the wind speed as necessary if other heights (usually 10 m) were used to replace missing 3-m values. In practice the difference in wind speeds was usually less than 10% with height for wind speeds greater than the 1-m s^{-1} threshold.

In our analysis of the wind data, we have taken a conservative approach, in that wind data that are suspect are marked as missing and excluded from all calculations. For example, any period greater than 24 h when wind speeds do not exceed the minimum measurable wind speed threshold (0.4 m s^{-1}) were marked as missing. While this may seem overly restrictive, close inspection of the data showed that episodes of minimum wind speed occurrence were bimodal in duration, either lasting less than 10 h or else persisting for several days. Typically these longer periods were abruptly terminated when wind speed and/or temperature increased enough to free the instruments. Wind direction values were treated in a similar manner, though there were many fewer periods in which the wind direction met the discard criterion of less than 10° variance in a 15-h period. In a few cases winds were very light at all sites simultaneously for several hours, and these data were not excluded. The net result of this quality filtering is most likely a bias for higher mean winds than in reality, since data excluded during periods of nonfunctional wind instrumentation are typically characterized by light winds. A less-frequent problem, partial icing of the anemometer resulting in lower wind speeds than occurred, is usually not detectable from the data alone and hence cannot be filtered.

In addition to being more difficult to measure than temperature, a climatology of the vector quantity wind is also much more challenging to present and discuss in a coherent fashion. Winds exhibit influences on several scales from synoptic pressure gradients to mesoscale influences (e.g., sea breezes and terrain-induced drainage flows) to very localized conditions (e.g., vegetation state and snow depth) and in turn can influence other environmental parameters in several ways (e.g., desiccation of vegetation and saltation, sublimation, and deposition of snow). Therefore, we present wind data in several ways, with an emphasis on how it may affect the redistribution of snow, the predominant surface cover for most of the year.

Table 4 contains monthly mean and record maximum wind speeds for each month. (The record wind speeds are, of course, the extrema for only the five-year period considered here.)

Wind speed is summarized in a manner analogous to our approach to temperature (Fig. 6), with 11-day running averages of daily mean wind speed and the maximum wind speed for each calendar day displayed. The dispersion (also known as the coefficient of variation, i.e., the standard deviation from the mean for each day, normalized by the mean) is also shown in Figure 6.

To summarize wind direction, data were binned by octant considering three threshold speeds: 1.0 , 4.5 , and 10 m s^{-1} (Fig. 7) and presented as wind roses. The 1.0-m s^{-1} threshold is indicative of all winds experienced at the site. (Winds less

Table 4. Monthly mean and record maximum wind speeds for each month.											
		BET		FRA		IMN		SAG		WKU	
		Wind speed (m s ⁻¹)	Day-year	Wind speed (m s ⁻¹)	Day-year	Wind speed (m s ⁻¹)	Day-year	Wind speed (m s ⁻¹)	Day-year	Wind speed (m s ⁻¹)	Day-year
EC											
Sep.	Record max.	16.4	28-95	18.5	26-96	14.2	25-98	12.5	27-95	13.4	26-94
	Monthly mean	5.4		4.5		3.6		4.0		3.8	
Oct.	Record max.	22.4	24-98	26.4	24-98	16.8	29-96	16.0	30-96	15.5	29-96
	Monthly mean	4.6		4.0		3.3		3.3		3.2	
Nov.	Record max.	21.2	3-94	20.3	19-98	19.0	21-96	17.4	11-97	16.4	30-95
	Monthly mean	5.1		4.1		4.2		4.3		3.3	
DC											
Dec.	Record max.	16.2	19-96	24.8	19-95	16.1	3-94	16.3	19-95	13.2	30-94
	Monthly mean	3.7		4.1		3.6		2.8		2.9	
Jan.	Record max.	17.3	13-97	22.4	13-97	16.5	17-94	15.8	29-96	16.8	24-95
	Monthly mean	5.0		5.0		3.9		3.2		3.2	
Feb.	Record max.	18.6	13-95	19.1	3-98	20.4	16-95	19.6	18-96	16.6	13-95
	Monthly mean	5.0		4.7		4.2		3.2		3.0	

than 1.0 m s⁻¹ in this data set tended to be rather variable in direction and did not contribute in a meaningful way to the direction statistics.) The 4.5-m s⁻¹ criterion, which we refer to as *moderate winds*, was chosen because it represents a median value for an accepted velocity that can readily mobilize fresh or unconsolidated snow in this area (Liston and Sturm 1998). The 10.0-m s⁻¹ threshold, which we refer to as *strong winds*, was chosen as a representative velocity capable of remobilizing older snow that is already somewhat consolidated or drifted. In addition to the five sites comprising the network, data from 1994–1997 for the Prudhoe Bay Airport (PRU) were also analyzed as a basis to compare the network data with a well-known coastal location. BET is about 25 km (15 miles)

Table 4 (cont.). Monthly mean and record maximum wind speed for each month.											
		BET		FRA		IMN		SAG		WKU	
		Wind speed (m s⁻¹)	Day-year	Wind speed (m s⁻¹)	Day-year	Wind speed (m s⁻¹)	Day-year	Wind speed (m s⁻¹)	Day-year	Wind speed (m s⁻¹)	Day-year
LC											
Mar.	Record max.	18.1	28-96	19.0	24-98	20.1	8-95	16.7	27-96	14.9	28-96
	Monthly mean	5.3		4.8		4.0		4.1		4.4	
Apr.	Record max.	14.9	19-96	18.2	27-98	13.7	25-95	18.9	15-97	19.3	15-97
	Monthly mean	4.0		3.5		3.7		3.1		3.4	
May	Record max.	17.7	19-98	19.7	21-98	12.6	4-95	14.5	19-98	13.0	20-98
	Monthly mean	5.6		5.0		3.8		3.9		3.9	
WS											
Jun.	Record max.	13.6	4-96	12.0	4-96	17.7	4-96	14.5	4-96	13.4	4-96
	Monthly mean	4.9		4.1		3.9		3.9		3.9	
Jul.	Record max.	14.9	23-98	12.2	24-98	15.1	27-96	13.9	27-96	11.4	27-96
	Monthly mean	5.1		4.3		4.0		4.2		4.0	
Aug.	Record max.	15.2	18-94	15.4	25-97	15.0	20-98	14.3	18-94	17.7	18-94
	Monthly mean	4.8		4.1		4.4		4.0		3.9	

west of PRU and FRA is approximately 40 km (25 miles) south of PRU. As PRU was operated as a privately contracted SAO (surface airways observation), little is known regarding the data quality and reporting practices for this station.*

Warm Season

During *WS* BET is clearly the windiest of the sites, in terms of both mean and peak daily winds, with the remainder of sites showing a great deal of similarity with regard to these mean quantities. BET also has the least dispersion,

* Personal communication with Ted Fathauer, National Weather Service, Fairbanks, AK.

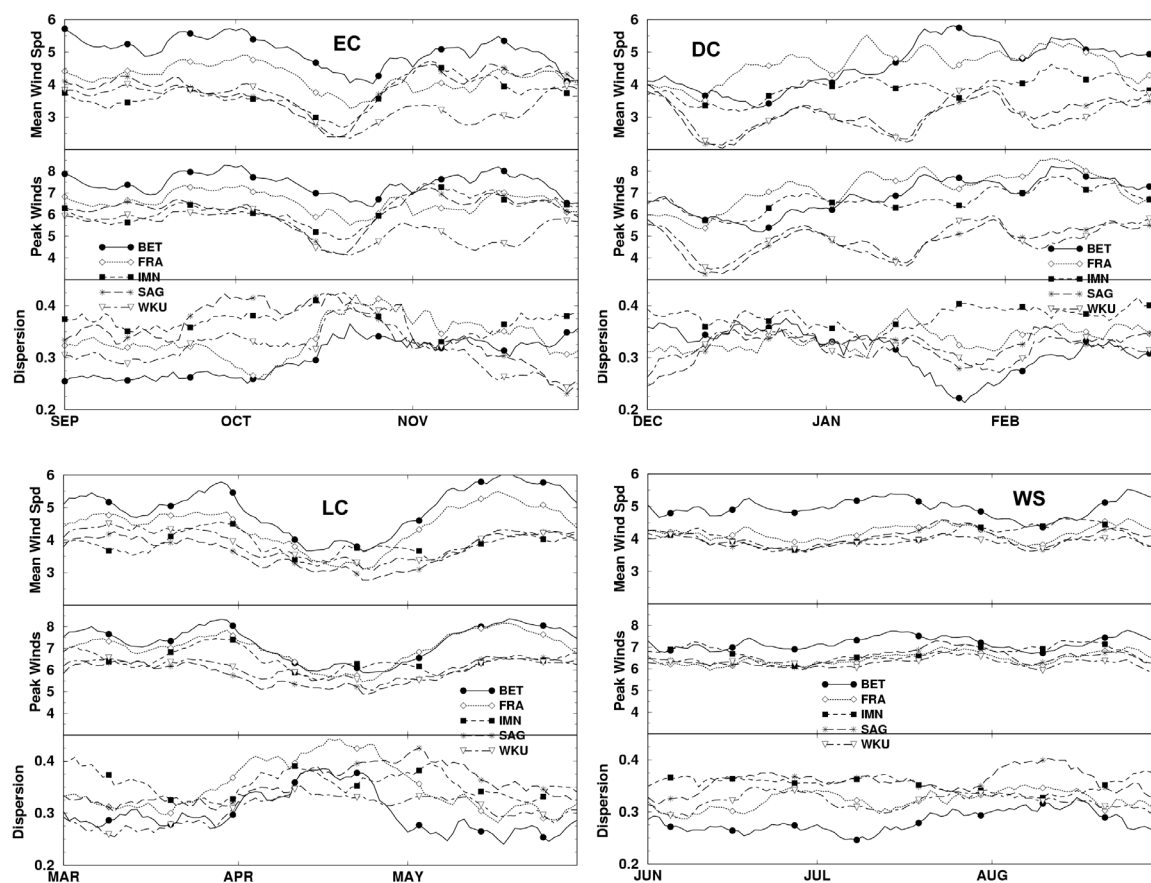
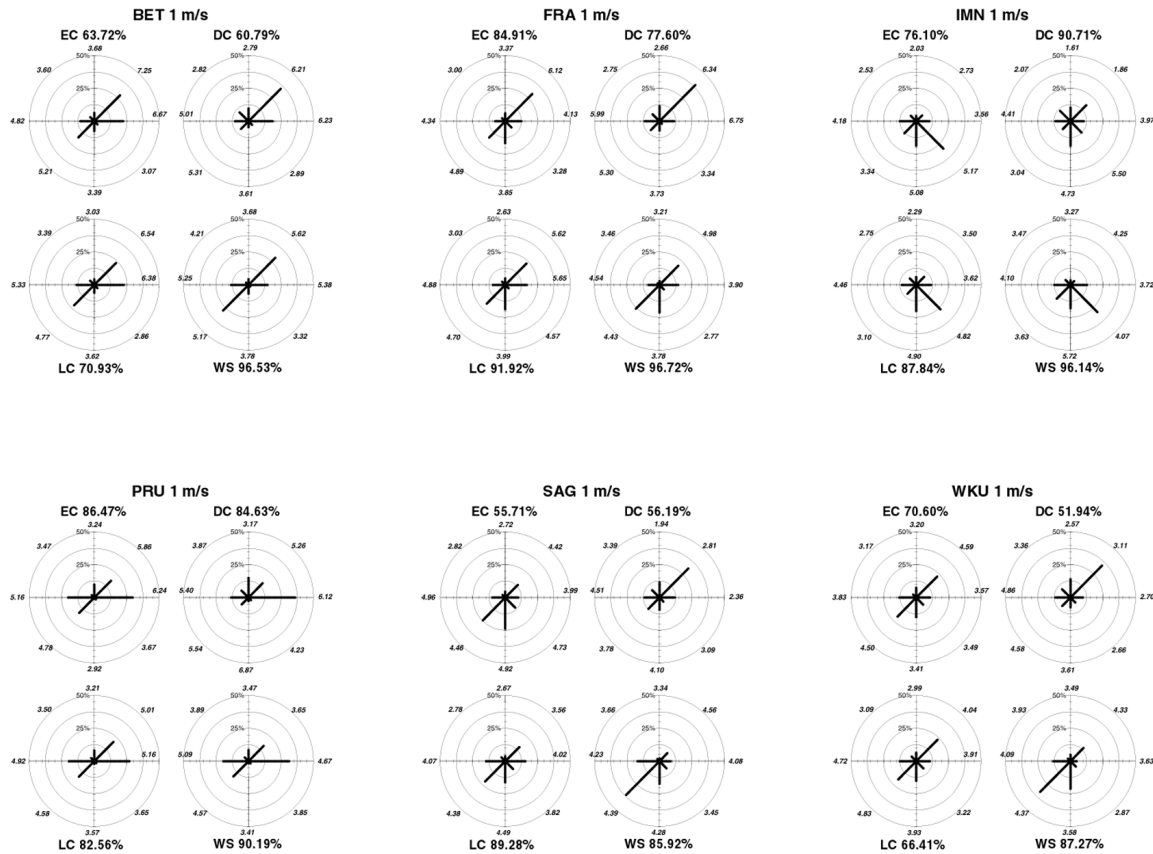


Figure 6. Eleven-day running averages of daily mean wind speed (m s^{-1}), average maximum wind speed (m s^{-1}), and dispersion (standard deviation normalized by mean) for each day for the four seasonal periods.

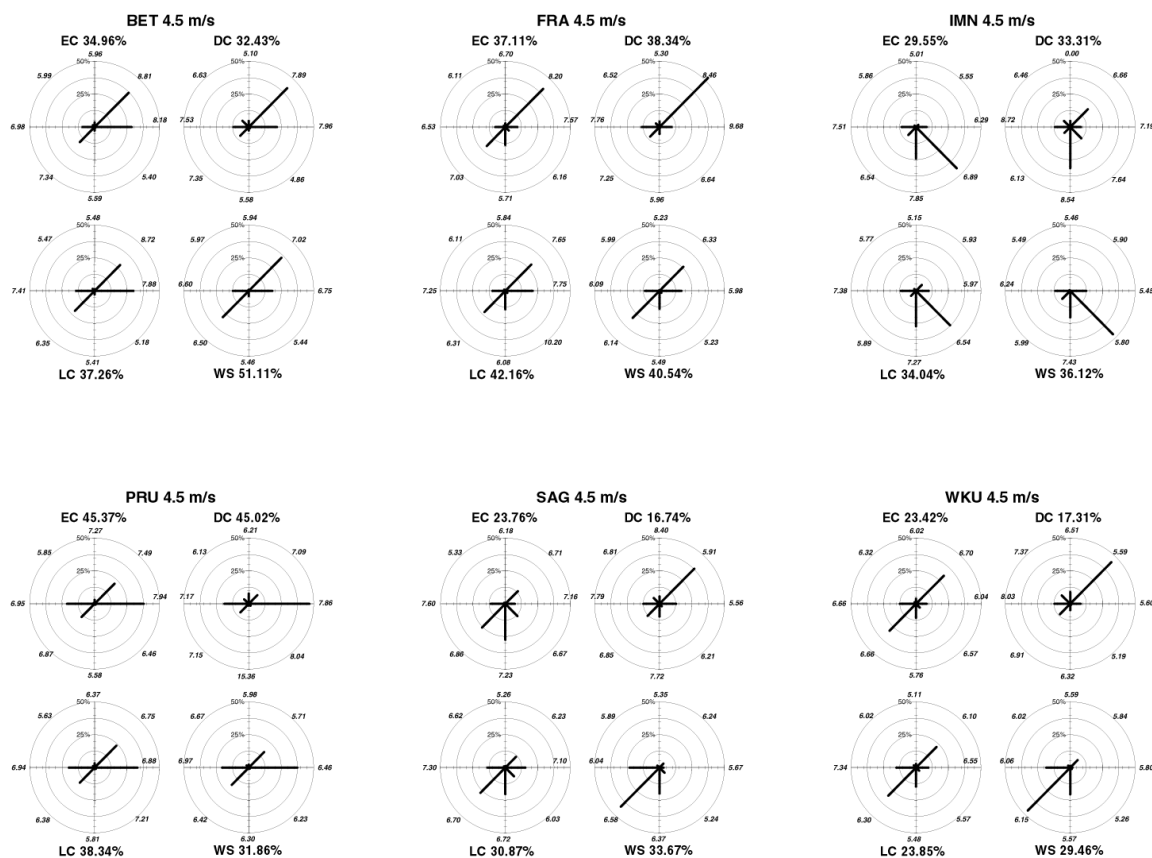
suggesting that this coastal location consistently has sustained winds in the neighborhood of 5 m s^{-1} . (Note that the dispersion is a measure of the variability of wind speed for all hours and years for a given day. This is a useful measure of variability for this study, but it is *not* an adequate measure of gustiness.) The wind rose diagrams (Fig. 7) show a bimodal distribution of wind direction, with almost equal likelihood of wind from the NE and SW. Prudhoe Bay (PRU), closer to the coast, shows a more even distribution of winds. BET is the only site indicating winds exceeding 4.5 m s^{-1} over 50% of the time, with the strong preference to NE and SW directions holding there as well. FRA shows a wind direction distribution similar to that of BET, which is not surprising considering that both are located on the coastal plain. BET and FRA show a large bias for strong winds from the E and NE.



a. Threshold minimum velocity of 1.0 m s^{-1} .

Figure 7. Seasonal wind rose diagrams for the five sites and for Prudhoe Bay using three threshold minimum velocities. The percentages listed for each season indicate the fractional number of all data points for that season which met or exceeded the threshold criterion. Labels for each octant indicate the mean wind speed for that direction for points that satisfied the threshold criterion. Missing data values were excluded from all calculations.

By contrast, foothills IMN sees almost *no* winds with a significant easterly component, while the intermediate SAG and WKU show a preponderance of SW winds. Probabilities for strong winds (defined here as winds $\geq 10 \text{ m s}^{-1}$) are less than 3% for any site in this season and less than 1% for the sites above the coastal plain, while the rare strong winds tend to be from S to W. Though not clearly evident from the plots shown here, during *WS* the upland sites, unlike BET and FRA, did not show wind regimes that persisted several days. Also, the strong wind events were frequently (though not always) quite transient and often occurred near the end of the diurnal heating cycle consistent with outflow from so-called air-mass convection. No winds exceeding 20 m s^{-1} were seen at any site



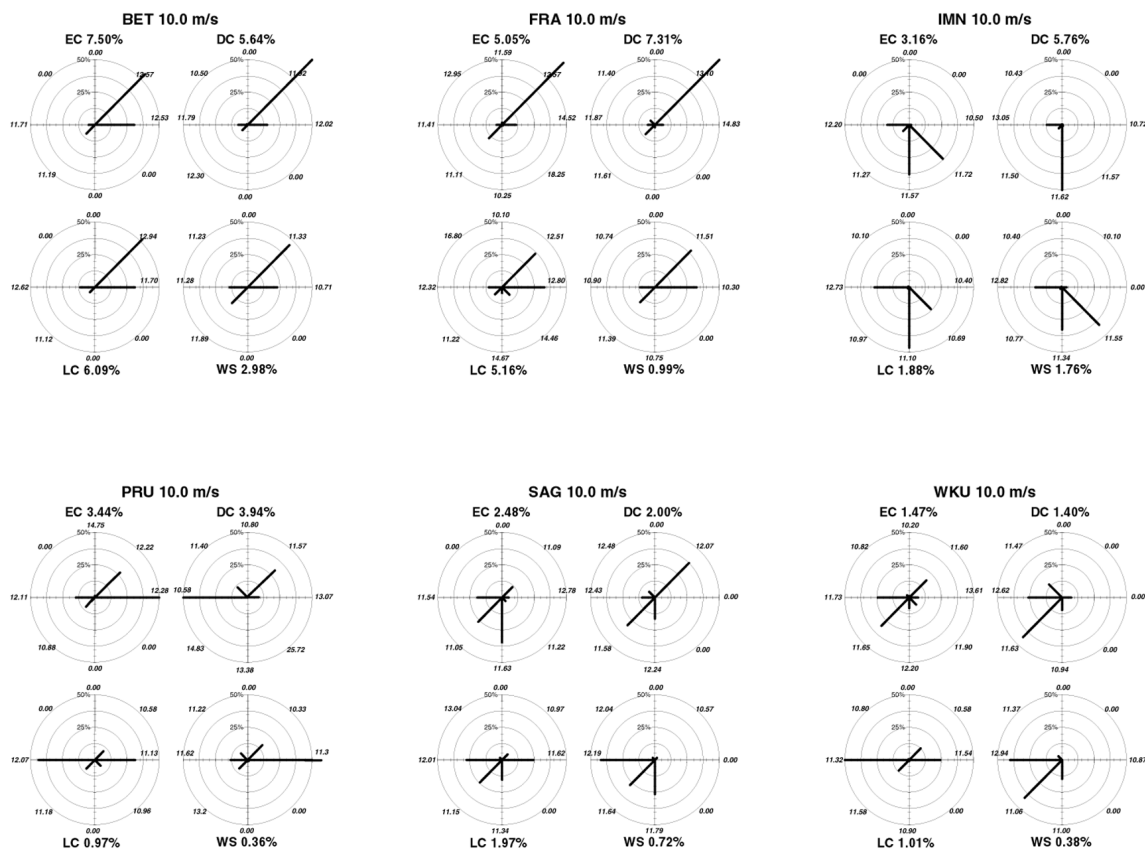
b. Threshold minimum velocity of 4.5 m s^{-1} .

Figure 7 (cont.). Seasonal wind rose diagrams for the five sites and for Prudhoe Bay using three threshold minimum velocities.

during *WS* for the period of record. It is also interesting to note that monthly records for this season seem to be consistent across the network, suggesting that strong wind events, though rare during this season, tend to arise from pressure gradients experienced across the entire transect.

Early Cold

During the first month or so of *EC*, wind speeds are similar to *WS*, with BET still clearly the windiest location. As the season progresses, wind speeds in the upland locations of IMN, SAG, and WKU tend to decrease, reaching a minimum in mid-October. This is significant because during this time the seasonal snow cover is beginning to accumulate. The light winds during these first “all-snow” precipitation events allow the new snow to accumulate in relatively pristine form,



c. Threshold minimum velocity of 10.0 m s⁻¹.

Figure 7 (cont.). Seasonal wind rose diagrams for the five sites and for Prudhoe Bay using three threshold minimum velocities.

resulting in low-density strata in the nascent snow cover. By the end of October the mean winds at SAG and IMN, the more exposed upland locations, increase. Also during this time, the dispersion in wind speed at BET increases and remains relatively high for the ensuing three-month period.

With the exception of PRU, the likelihood of moderate to strong winds (4.5 m s⁻¹) decreases from *WS* to *EC* for all sites and is less than 30% for the upland locations. Both BET and FRA continue to show a bias for winds from the NE and SW, but winds with an easterly component become more common, and as wind speeds increase, so does the likelihood for E to NE winds. *EC* is the most common season for BET to receive strong winds. By contrast, winds at the intermediate locations of SAG and WKU show less directional preference, while winds with a northerly component are quite rare at IMN, where the majority of winds

come from the SE to E. The strongest wind of record occurred in October at the coastal plain sites, with the October record at FRA of 26.4 m s^{-1} (59 mph) being the highest wind speed recorded on the network during this five-year period. In summary, *EC* in the uplands results in low wind conditions, while the coast and coastal plains are windier.

Deep Cold

Mean wind regimes seem to be best established during *DC*, with the coastal plains sites BET and FRA the windiest, the uplands between the coastal plains and the Brooks Range foothills having seasonally the weakest winds found in the record, and foothills IMN falling between these two regions. The same regimes generally exist for peak winds, except IMN, which has sufficient windy episodes to bring the smoothed results similar to those found along the coast.

The coastal locations show a gradual increase in mean winds from a minimum in early December and a strongly preferred wind direction from the NE to E. This preference becomes even more pronounced when considering the strongest winds, which are almost exclusively northeasterlies. It is curious that PRU shows a decided preference for strong *westerlies*. (It should be noted that the PRU record for these years shows much less likelihood for winds $\geq 10.0 \text{ m s}^{-1}$ than the other coastal stations. We were unable to determine a reason for this discrepancy.)

Considering the direction for winds of all speeds, IMN has a fairly uniform distribution throughout the octants. As wind speed increases, however, the slight preference for southerlies becomes more and more pronounced, almost completely dominating the record for the strongest winds. *DC* is also the most likely season for winds at IMN to exceed the 10-m s^{-1} threshold. IMN's location on the northern edge of the deep Atigun Valley makes it ideally situated to experience strong southerly channeled flow during periods when interior Alaska is experiencing high surface pressure. Such gap flows (Reed 1931, Scorer 1952) are commonly observed in other regions of Alaska where mountains separate disparate air masses (e.g., Macklin et al. 1990, Bond and Macklin 1993). Since *DC* is also the most likely time of year for the existence of the stability criteria that support mountain waves and the development of supercritical flow (e.g., Durran 1986, Klemp and Lilly 1975) associated with downslope windstorms as well as those favoring gap flows, it seems likely that these southerly wind events are inherently mesoscale in nature and likely confined to a region near and to the lee of the higher terrain of the Brooks Range. At lower wind speeds the intermediate sites of SAG and WKU both show the preference for NE winds seen on the coastal plain. However, for the strongest winds the probability of winds from the

S to W increases substantially, especially at WKU, where NE winds exceeding 10 m s^{-1} are nonexistent in this record.

Late Cold

For March, the first month of *LC*, wind speeds are similar to the last half of February. A similar tendency is seen in mean temperatures in the first half of March, suggesting that, though the daily increase in insolation is near its peak at this time, winter weather continues to dominate the Arctic Slope with an iron grip. In April, one of the driest and least stormy months of the year in northern Alaska, the coastal plain winds (BET and FRA) decrease substantially for a short period, but by May coastal winds increase to the point that BET sees its largest mean wind speeds of the year. At IMN wind speeds slowly decline throughout *LC*, and by late May the wind speeds at the three most inland locations are similar. Monthly record wind speeds at IMN also decrease through the season, with the 12.6 m s^{-1} for May the lowest for this station. The likelihood for strong winds at the sites for *LC* is similar to that in *DC* with the exception of IMN, where the conditions necessary for terrain-enhanced winds are less likely to occur.

5 SNOW COVER

In addition to the meteorological data from the five tower locations, annual surveys of the snow cover along a transect down the Kuparuk River (Fig. 8) were conducted during the period covered by this report. Snow pits were dug and examined in 30 locations between the Brooks Range and the coast. Acting as an integrator of cold-season weather events, these “snow-pit traverses” present a complementary picture of the cold-season climate in the Arctic Slope, showing periods of both low and high wind, as well as extended periods of cold.

Solid precipitation, wind, and air temperature are the three main weather controls on the development of a snow cover. Precipitation and wind are primary controls, determining the depth, density, and characteristics of the snow cover.

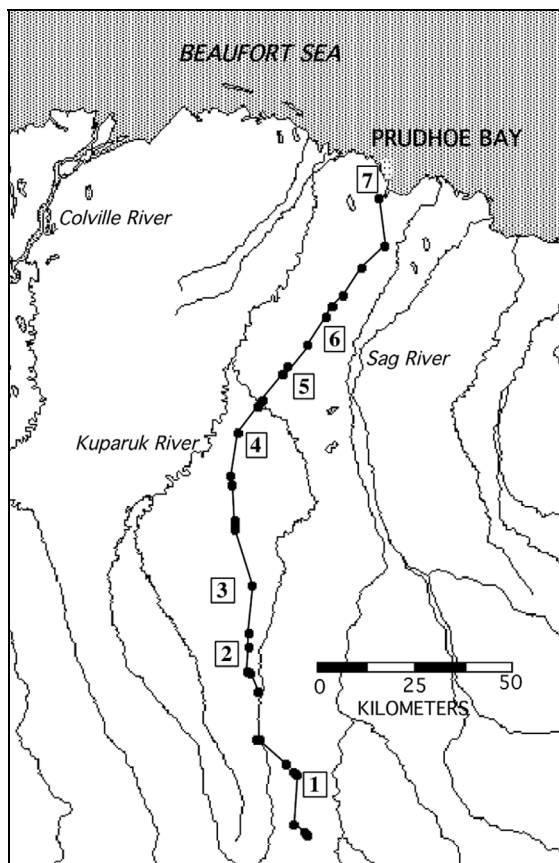


Figure 8. Line of the winter traverses in the Kuparuk Basin and the locations of the seven snow pits shown in Figure 9.

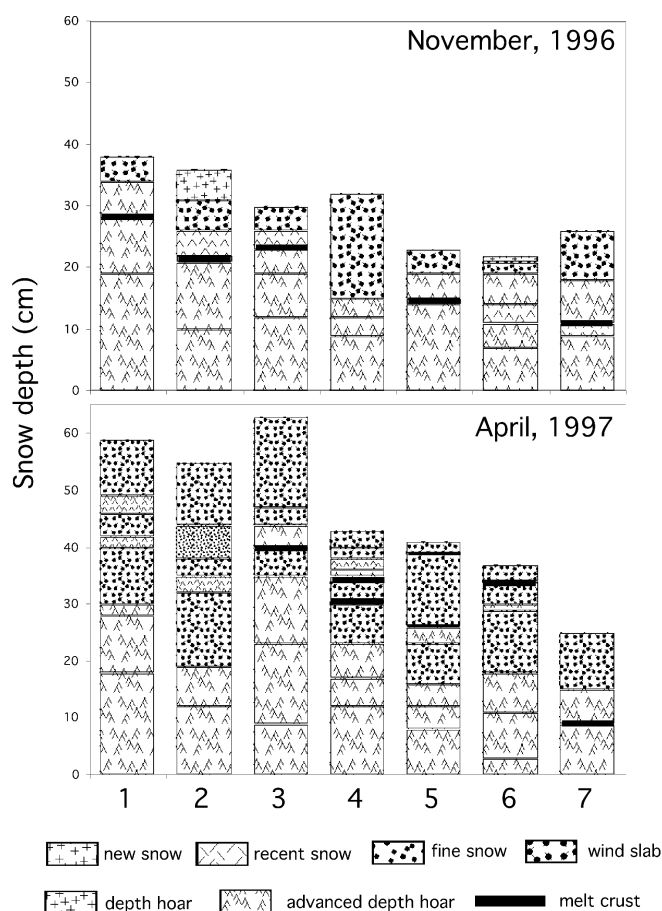


Figure 9. Stratigraphy for the Kuparuk Basin, winter of 1996-97. The locations are shown on Figure 8. Symbols are from the *International Classification for Snow on the Ground* (Colbeck et al. 1992). Note the two to four layers of depth hoar at the base of the snow, capped in most cases by wind slabs and/or melt crusts.

Because most of the vegetation of the Kuparuk consists of grasses, sedges, and low shrubs, wind transport of snow is common (Benson and Sturm 1993). Temperature plays a more subtle but important role, determining the extent of depth hoar growth (a coarse-textured type of snow at the base of the snowpack formed through metamorphic processes), the hardness of slabs deposited by the wind, and the number and spatial continuity of melt crusts. While the basic functional relationships between the weather and the snow cover stratigraphy are known (i.e., longer and more-intense snowfalls produce thicker snow layers), an infinite number of weather sequences are possible. The same set of weather events com-

ing in a different sequence can produce a wide range of snowpack characteristics, and these have a commensurately wide range of impacts on the environment and man.

Surprisingly, but for the same reasons that the winter climate of the Kuparuk tends to fall within a narrow range, the sequence of weather events observed along the Kuparuk River is sufficiently similar from year to year to produce a similar snowpack. As a consequence, stratigraphic and textural characteristics of the snow tend to be similar from one year to the next—sometimes strikingly similar—even though the potential exists for the snow cover to exhibit a much wider range of characteristics.

Figure 9 shows the snow stratigraphy of the Kuparuk Basin for early and late winter. The total snow depth and the layer thickness varies considerably, but several features repeat from one location to another:

- Two to four relatively thick layers of depth hoar are found at the base of the pack in most places,
- The depth hoar is capped by a thick wind slab that is in place by late November,
- Relatively few layers are added between December and April, and
- Overall, no more than eight distinct layers comprise the total pack.

These salient characteristics are not unique to the winter of 1996–97. For all winters for which we have snow pit data (1989–1997), similar features were observed. Figure 10 shows the snow cover stratigraphy near the end of the cold season for the four winters of 1995–1998. Despite differences in depths and layer thickness, overall the same characteristics appear.

The snow characteristics are linked directly to the climate and weather events. Depth hoar is produced when relatively fluffy, low-density snow is subjected to strong temperature gradients for days to weeks (Akitaya 1974, Trabant and Benson 1972, Sturm 1991). Virtually every winter on the Arctic Slope, this condition is met. Further, the formation of depth hoar is enhanced because early-season snowfalls tend to be light and fluffy, falling with little wind (Fig. 6), though the reason for this low wind regime and associated snowfall is not well understood. Wind slabs are produced by winds in excess of $4.5\text{--}6\text{ m s}^{-1}$, provided there is sufficient snow available for transport (Kuz'min 1963). The snow pit data strongly suggest that the conditions needed to produce these strong winds tend not to materialize until the end of *EC*, when much of the season's snow cover has already been deposited. In the following pages we will examine some of these characteristics and the recurrent weather events that give rise to them, and we will explain them where possible.

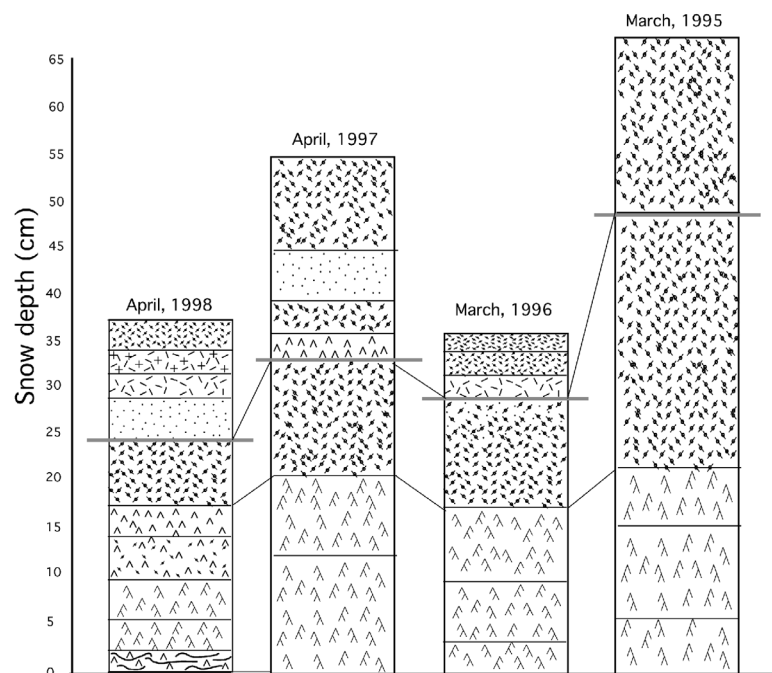


Figure 10. Snow stratigraphy from the Kuparuk Basin for the winters of 1994-95 through 1997-98, showing similar sequences of layers at the same site. The horizontal lines indicate the fall capping event, in most cases a dense wind slab and melt crust.

6 DISCUSSION

The results presented here suggest that while significant differences in climate exist latitudinally and altitudinally across the Kuparuk Basin, the climate overall is quite similar. The greatest departures from a mean annual cycle for the Kuparuk Basin result from mesoscale factors related more to geography than to meteorology—the coastal proximity of BET with its seasonal maritime influence and the foothills location of IMN with its higher elevation and susceptibility to orographically enhanced circulations. Indeed there is even a seasonality to the maritime influence at BET in that once the Beaufort Sea is largely ice-covered, net heat and moisture fluxes at the ocean–atmosphere boundary become quite small and BET becomes more continental in nature.

The results presented in the previous section are the result of smoothed averages, permitting comparison of the climate states among the sites. To compare specific meteorological events, it is instructive to consider an *unsmoothed* time series of daily mean temperatures for a particular season. Figure 11 shows such a plot for the period from Oct. 1997 through Jan. 1998. With the notable exception of IMN (dotted line) from December onward, the temperature trends to first order are generally quite similar from station to station, suggesting that the individual synoptic events that bring substantive temperature changes in this period occur across the entire network. The heavy curve in Figure 11 represents the average of BET, FRA, SAG, and WKU, with IMN being excluded from the calculation due to its quite different cold season thermal nature as discussed later.

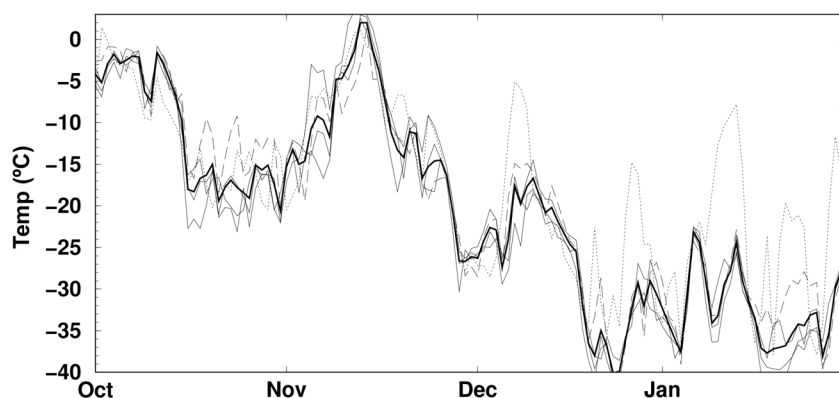


Figure 11. Daily mean temperatures at the five sites for Oct. 1997 through Jan. 1998. The bold line represents the average temperature of BET, FRA, SAG, and WKU; the dotted line represents the temperature at foothills IMN; and the dashed line represents coastal BET. No temporal smoothing was performed here.

The greatest variation among the four stations contributing to the mean curve in Figure 11 occurs from mid-October to mid-November, when BET (long dashed line) experiences locally strong but shallow boundary layer modification due to open ocean heat fluxes in off-ice (on-shore) flow (e.g., Olsson and Harrington 2000). During these conditions, typified by cold advection with strong NE winds, BET may actually *warm* temporarily due to maritime modification, while more inland and higher elevation locations, most notably WKU, experience falling temperatures. This effect at BET disappears with the onset of *DC* as ice is both pushed by the prevailing northeasterly winds against the Arctic coast and simultaneously grows in situ.

Also, with the onset of *DC*, the temperature relationship between IMN and the other sites becomes decoupled. During this season IMN experiences very dramatic warming events echoed weakly, if at all, by the mean curve. These IMN warming events are typically concurrent with strong southerly flow not seen at the other locations, again suggestive of compressional warming in downslope conditions. Temperatures drop rapidly once cross-barrier pressure gradients decrease and NE flow is again established.

Another notable feature in Figure 11 is the dramatic warm-up at all sites in the second week of November. All of the non-coastal sites experienced warming to above-freezing daily mean temperatures, with BET showing a weaker but still discernible warming as well. Further investigation shows that similar, if not as spectacular, warming events occur at least once during the latter third of *EC* at most sites for each of the years in the record. The timing of this warm-up, sometime between julian days 303 and 323 (the last week in October to the third week in November), is consistent enough and strong enough to appear clearly in Figure 3, having survived both the multiple year and running average operations.

Figure 12 shows the unfiltered time series of daily mean temperature during the latter half of *EC* at SAG for each of the five years of record. All years show one, and sometimes two, strong warming events, with many lasting for several days. [Though 1994 does not show a daily mean temperature that exceeds freezing, there were several hours spanning two days that the temperature was within 2°C of freezing. The fact that many nearby locations had experienced above-freezing temperatures was attested to by the widespread very hard and distinct snow stratum still quite evident at the onset of the melt season (Olsson et al. 2000)]. From the context of the cold-season snow cover, this annual event is significant. The resulting higher-density wind slab and melt crust (Fig. 10) acts as a “capping layer” that has structural, mechanical, and thermodynamic implications as the snow cover accumulates and evolves through the next six months.

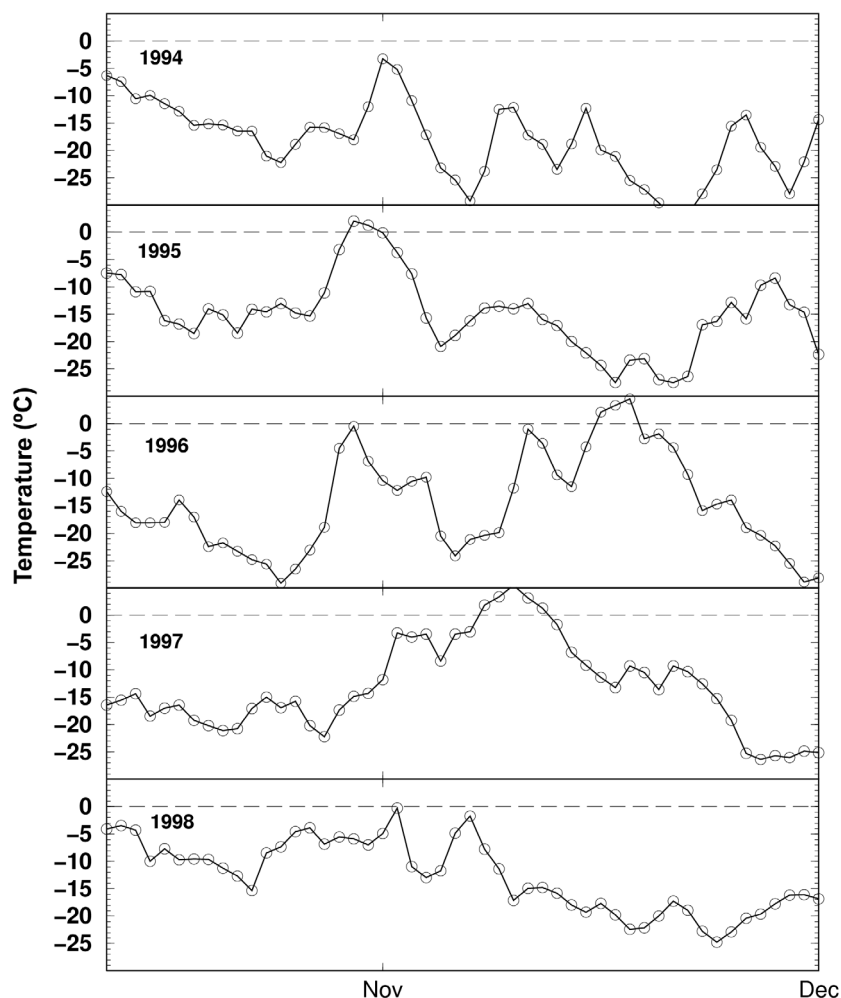
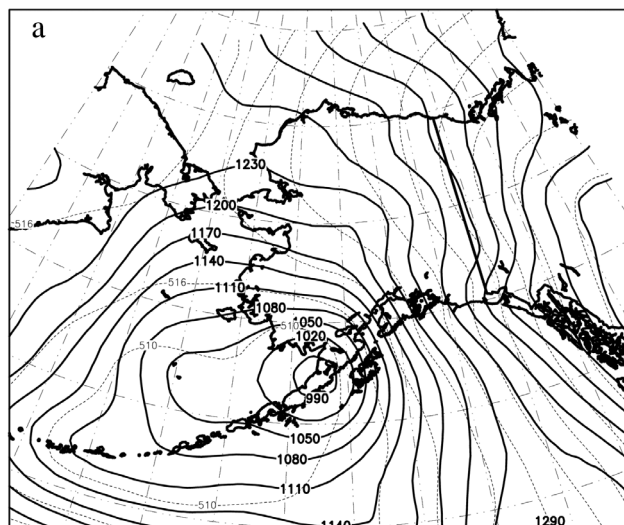
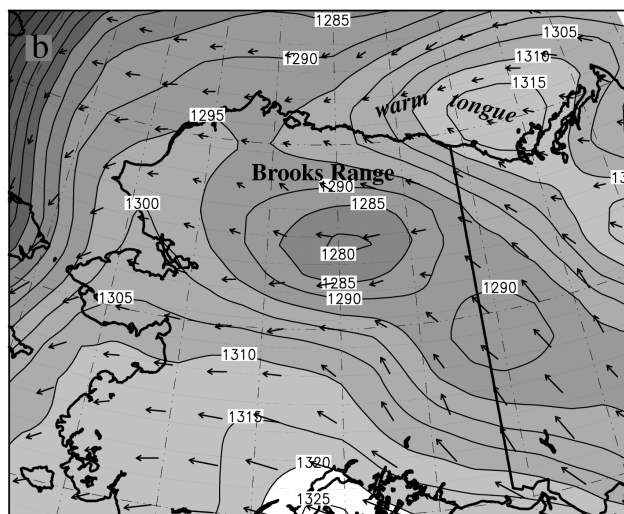


Figure 12. Daily mean temperatures at SAG for all five years of record during the latter half of *EC*, showing warming/thawing events (note the 0°C isotherm).

The *EC* warming event of 1997 (Fig. 12) had both the greatest duration and warmest temperatures of all these events in the period of record, with above-freezing daily mean temperatures from 8 to 13 November. Figure 13a shows the 12 UTC, 8 Nov., 1997 850-mb (heavy solid) and 500-mb (light dashed) geopotential height over Alaska and the North Pacific, nearly coincident with the first above-freezing temperatures at SAG. The lower troposphere (850-mb) height pattern was dominated by a deep extratropical low-pressure system in the Bering Sea centered over the Alaskan Peninsula–Bristol Bay region. The 850-mb geostrophic flow over eastern Alaska at this time was nearly meridional, with a strong W–E pressure gradient producing southerly winds. The midtropospheric



a. 850-mb geopotential height (meters, solid heavy contour) and 500-mb geopotential height (dekameters, dashed light contour).



b. 1000–850-mb layer thickness (m) and layer mean winds. Warmest areas have lightest shading (greater thickness) and coolest areas have darkest shading (smaller thickness).

Figure 13. Synoptic conditions for 12 UTC, 8 Nov. 1997. Data were derived from the initial analyses of the U.S. National Meteorological Center (NMC) ETA numerical forecast model.

500-mb pattern was quite similar, with a cutoff low centered in roughly the same location. This configuration had existed for the previous few days, with the nearly stationary and vertically stacked cyclonic structure drifting very slowly eastward.

Figure 13b shows the 1000–850-mb thickness (proportional to the mean temperature in this layer) and layer-mean winds for the same time. A tongue of warm air extended out of western Canada over the eastern Arctic Slope of Alaska and Beaufort Sea. The product of several days of previous warm advection from the circulation around the Bering Sea system, this warm tongue continued to be reinforced by warm surges in the SE flow for the next few days, resulting in the warming event seen in Figure 12. Interestingly the cold pocket over the Upper Yukon Basin south of the Brooks Range persisted throughout the warming event experienced on the Arctic Slope. Further analysis (Olsson 2001) suggests that this pattern is broadly typical of those producing the capping events seen in the temperature and snow cover records. Examination of the full record for SAG (1986–1998) shows late *EC* warming events occurred at least once in eight of the twelve years where the data record was complete for *EC*.

7 SUMMARY AND CONCLUSIONS

Presented here is a summary of five years of nearly continuous hourly surface meteorological measurements at five locations in the Kuparuk Basin on the Arctic Slope of Alaska, extending from near the coast to the foothills of the Brooks Range and spanning an environment from the coastal plain near sea level to the uplands region with elevations over 900 m. The data were analyzed and discussed largely in the context of running and seasonal averages and in the context of four quarter-year seasons. As well as considering temperature trends, this study developed a wind climatology for the five stations, with a thresholding scheme based on the potential of windiness to impact on the snow cover, the dominant ground cover for over eight months of the year.

A picture emerges of the Arctic Slope as a region dominated by subfreezing temperatures for most of the annual cycle; the brief warm and snow-free season in June, July, and August is even shorter than that of Arctic Alaska south of the Brooks Range. The five sites in the 200-km-long transect, while exhibiting some degree of variation from what might be considered an “Arctic-Slope-wide” climatology, still showed a good deal of similarity in both seasonal variation and meteorological forcing on a time scale of a few days.

Overall, the Arctic Slope is typified by almost constant light to moderate winds. Contrary to the common image of the Arctic Slope as a vast windswept region dominated by blizzard conditions, strong winds—those capable of remobilizing snow that has already been consolidated into the snow cover—are actually rather infrequent. The more coastal locations tend to experience NE and SW winds. Farther inland the wind direction becomes more variable. The southernmost site, IMN (Imnaviat Creek), typically has some southerly component to the wind in all seasons, in large measure because of its proximity to the Brooks Range.

Southerly winds at IMN are favored by a regional terrain configuration that can induce drainage flows as well as gap and wave-enhanced circulations. IMN tends to have episodic southerly strong wind events in Deep Cold (*DC*) and is also frequently decoupled at this time from the thermal regime of the lower sites. The overall variability of wind conditions during the cold season, coupled with the thermal influence of open water along the coast in October and November, suggests that caution should be used in extrapolating the conditions found along the coast (where the best long-term climate records exist) to the inland and upland locations of the Arctic Slope.

A notable feature of the annual temperature curve is a pronounced “bump” or short-term warming trend in the otherwise steady decline of temperatures across

the Arctic Slope during Early Cold (*EC*). Focusing on one representative site, it is found that this bump occurs to some extent in all years of the period of record. These annual late *EC* warming events result from nearly stationary or slowly propagating and vertically stacked low-pressure systems over southern Alaska that advect warm air northward to the Arctic Coast. Such events typically bring from several hours to several days of above-freezing temperatures to the Arctic Slope after a significant amount of snow has accumulated and produce a high-density stratum in the snow cover, capping the accumulated *EC* snowpack and protecting it from subsequent wind erosion.

The observed patterns of temperature and wind have important ramifications for the winter snow cover of the Kuparuk Basin, which plays a central role in the hydrology, ecology, and surface energy balance of the region. The seasonal differences in the temperature patterns between the Brooks Range and the coast explain why the snow cover forms first in the foothills and last near the coast, the presence of an ice-free ocean moderating early-winter coastal temperatures sufficiently to retard the formation of the snowpack. Similarly the occurrence of above-freezing temperatures near the Brooks Range more than a month before similar temperatures are experienced at the coast leads to a disappearance of snow in the south considerably earlier than at the coast.

The low wind speeds experienced across the network in October and November, particularly in the southern parts of the region, help to explain the presence of thick but low-density layers in the base of the pack throughout this area. These layers show signs of having been deposited in low wind conditions, and their presence is one of the reasons the pack is such a good insulator, retarding heat loss from the ground throughout the winter. The insulating value of these basal layers is greatly enhanced when they metamorphose into depth hoar, which has the lowest thermal conductivity of any type of snow (Sturm and Johnson 1992, Sturm et al. 1997). This metamorphic transition is virtually assured by the cold winter climate and low winter precipitation values of the Arctic Slope. What is surprising is that these low-density snow layers, which are at risk of erosion during later high-wind events, are not eroded most of the time. Their persistence can be in part ascribed to the unusual thaw observed each winter in November as well as the generally low winds observed in much of *EC*. The thaw has been observed to create an icy cap at the top of the snowpack that resists wind erosion. We are unsure what happens in years when there is no thaw, but we note that a close and critical connection exists between the snow cover and the weather. This connection relies not only on the broader features of the weather but may depend also on some of the subtler aspects.

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